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A hydraulic box model study of the Mediterranean response to postglacial sea-level rise

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[1] This paper quantifies the role of changing sea level in affecting Mediterranean stratification using a new model of the strait-basin system, which allows for explicit time dependence. During the last deglaciation, Fairbank's Meltwater Peak 1B leads to rapidly rising global sea levels and a freshening of the oceans and hence increasing influx of ever freshening Atlantic water into the Mediterranean basin. Owing to long residence times, the salinity of the deep and intermediate water reservoirs would have decreased more slowly, so that the basin was more stably stratified for a period, with possible implications for sapropel formation. Our model is used to estimate the size of this reservoir effect in comparison to the freshwater influx from the opening of the Black Sea at the Bosphorus. It is found that this mechanism affects the stratification to a similar degree as the Black Sea opening. Other climatic mechanisms have also affected the freshwater budget of the Mediterranean at this time, but their sizes are more difficult to quantify. *INDEX TERMS:* 4243 Oceanography: General: Marginal and semienclosed seas; 4556 Oceanography: Physical: Sea level variations; 4255 Oceanography: General: Numerical modeling; 4267 Oceanography: General: Paleoceanography; *KEYWORDS:* Mediterranean Sea, Strait of Gibraltar, Holocene, sea level, hydraulic control, box model

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1. Introduction

[2] Owing to its antiestuarine circulation, the present-day Mediterranean is well ventilated throughout the water column, and the photic zone is nutrient-depleted, resulting in low primary productivity, particularly in the east. However, sediment cores have revealed more than 150 sapropel layers in sediments since the middle Miocene. A sapropel is defined as a discrete sediment bed containing at least 2% organic carbon by weight, while sapropelitic layers contain between 0.5% and 2% organic carbon [*Aksu et al.*, 1995]. Generally, sapropels indicate that the production of organic carbon exceeded the oxidation of dead organic matter, so that part of the carbon could be deposited. This might happen if remineralization were arrested in the deep waters due to reduced ventilation and oxygen availability, or if surface primary productivity were extremely high. These mechanisms are not mutually exclusive since reduced convection can lead to upwelling of nutrient rich waters and increased productivity [*Rohling and Gieskes*, 1989; *Kemp et al.*, 1999].

[3] The youngest sapropel layer S1 was deposited during the Holocene between 9–6 ka BP, with the onset in the Adriatic Sea rather later than in the Aegean Sea [*Aksu et al.*, 1995; *Mercone et al.*, 2001], at a time when global climatic changes were occurring, including the recent deglaciation. A number of mechanisms have been proposed to explain the occurrence of S1, mainly invoking stagnation through the influx to the eastern Mediterranean of large amounts of comparatively fresh water at the surface. It is not the aim of this paper to discuss all possible sources which might have freshened the Mediterranean but rather to attempt to quantify and compare the mechanisms directly associated with postglacial sea-level rise where the size of the freshwater input can be reasonably well quantified.

[4] The main focus is the changing freshwater inflow through the Strait of Gibraltar. The period since the Last Glacial Maximum (LGM) is characterized by a retreat of glaciers and melting of the ice caps, leading to a total sea-level rise of 120 m. *Fairbanks* [1989] notes that sea-level rise was most rapid in two periods: The first Meltwater Peak ("Meltwater Peak 1A") is centered around 12000 years BP and ends at the beginning of the colder Younger Dryas chronozone, while the second ("Meltwater Peak 1B") occurs after the Younger Dryas, centered around 9500 years BP. Figure 1 shows the sequence of some relevant events, and Figure 2 shows the sea-level curve used in this study [after *Fairbanks*, 1989].

[5] The Strait of Gibraltar limits the water exchange with the Atlantic, so that net evaporation leads to higher salinity in the Mediterranean Sea [e. g., *Bryden and Stommel*, 1984]. This limiting effect was more pronounced at times of lower sea level when the strait was narrower and shallower. As the sea level rises, the surface inflow of Atlantic water increases, but the properties of the deep bulk of the Mediterranean water can change only slowly, leading to a

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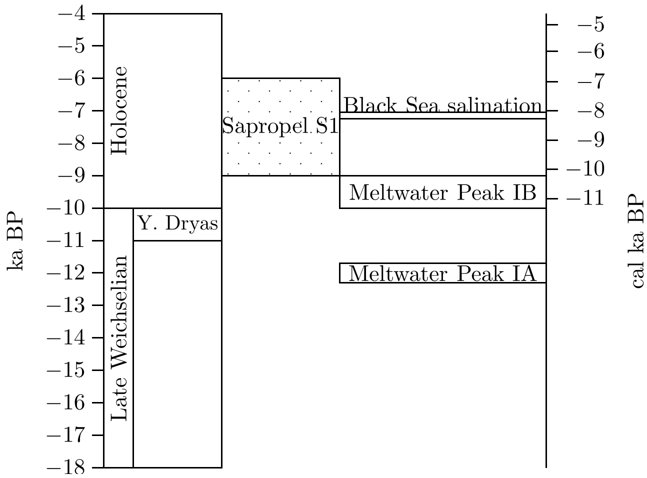

Figure 1. Stratigraphy and some relevant events since the Last Glacial Maximum. Left axis in uncalibrated radiocarbon ages, right axis in calibrated years (the calibration is not well established beyond 10 ka BP).

transient increase in stratification at times when the sea-level change is sufficiently rapid [*Rohling*, 1994]. This lag in the deep water properties can be called the reservoir effect. This mechanism occurs independently of other changes in the freshwater budget, although the inflow salinity would also have decreased due to the release of freshwater from the cryosphere to the Atlantic.

[6] The other effect which is directly related to the rising sea level is the opening of the connection to the Black Sea at the Bosphorus. The Black Sea was a freshwater lake ("Black Lake") in the late Pleistocene, and as the connection was established a sudden influx of freshwater to the Mediterranean would have occurred. There are two possible scenarios for the opening of the Bosphorus. The gradual inflow scenario [e. g., *Lane-Serff et al.*, 1997, and references therein] assumes that the Black Lake always had a freshwater outflow to the Mediterranean and that therefore the freshwater supply to the Aegean may have been similar to present day before the Bosphorus was opened. When the sea level had risen sufficiently above sill depth (at 40–60 m below present sea level), Mediterranean water started to flow in at the bottom, converting the Black Lake into a brackish Black Sea and establishing the present two layer system. A considerably increased freshwater outflow would have accompanied the bottom inflow for two or three millennia until the Black Sea was filled with saline water [*Lane-Serff et al.*, 1997]. This outflow would have increased the stratification of the eastern Mediterranean helping to reduce deep water ventilation. The increased inflow of fresh surface water would have been equivalent to an increase in net precipitation over the whole Mediterranean of 13 to 40 cm/year depending on the time period of the exchange.

[7] An alternative scenario by *Ryan et al.* [1997] suggests that the connection between the Black Lake and the Mediterranean was lost completely in the late Pleistocene, and the Black Lake level dropped to approximately 120 m below the present level. The opening to the Mediterranean would then have resulted in a catastrophic waterfall and subsequent drowning of the Black Sea shelf, with the present-day two-layer exchange established within only a few years. An unconformity in the Black Sea has been detected and dated to 7150 ± 100 years BP. In this scenario, the freshwater input at the Bosphorus is negligible before 7150 years BP, and increases quickly to above the present value as the Black Lake freshwater is lost, and then returns to present-day values. The main difference

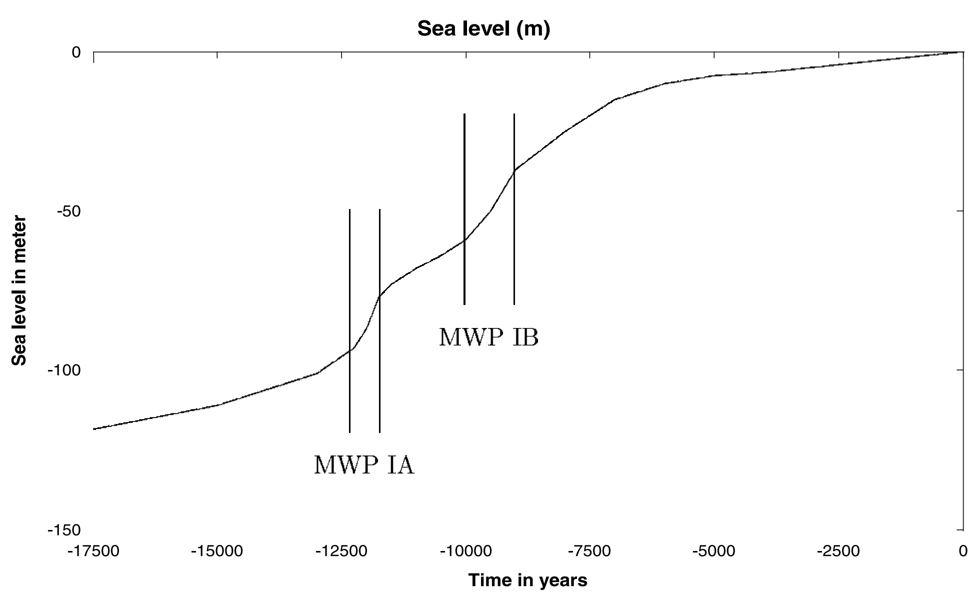

Figure 2. Sea-level change during the Holocene, showing the two Meltwater Peaks IA and IB after *Fairbanks* [1989].

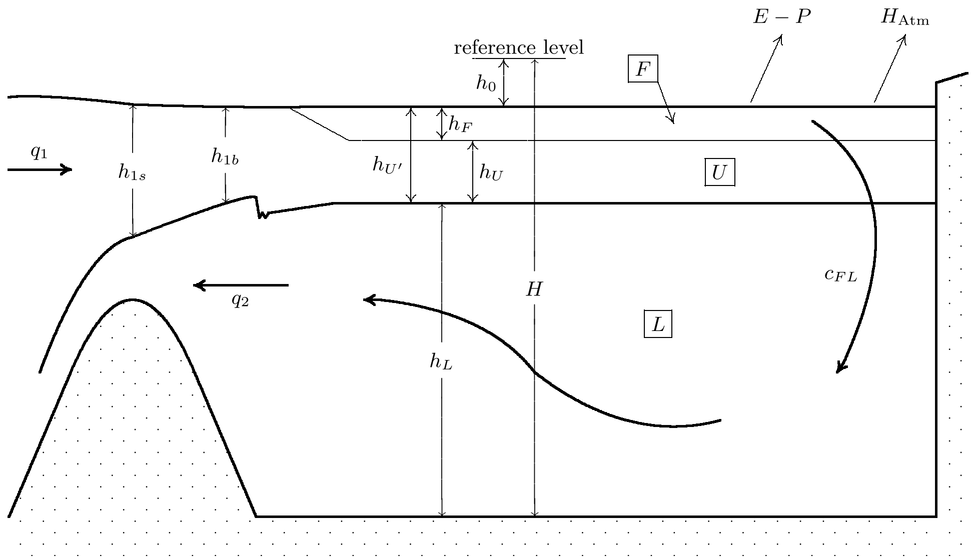

Figure 3. HYCOBOX, the hydraulically controlled box model, consists of a basin with three boxes F (water formation box), U (upper layer box) and L (lower layer box), connected to an infinite reservoir through a hydraulically controlled strait. Only the water formation box F interacts with the atmosphere. Boxes F and U together can be regarded as an effective upper layer U'.

between these scenarios for the Mediterranean is that the volume of freshwater flushed out is about 15% smaller in the catastrophic scenario, and the preopening freshwater flux to the Aegean would be below present day, so that a permanent increase in freshwater inflow to the Mediterranean persists after the opening due to the draining of Black Sea rivers.

[8] Other climatic events are also known to have affected the freshwater budget of the basin during this period. An increased freshwater input from the Nile due to monsoon intensification around 10 to 6 ka BP [*Rossignol-Strick*, 1985; *Rohling*, 1999], and increased rainfall in the northern borderlands of the eastern Mediterranean [*Rohling and Hilgen*, 1991], are two such mechanisms. However, we will not attempt to quantify these other effects, as the objective of this paper is to quantify effects directly associated with the rising sea level and in particular to show that the reservoir effect alone is important and may have had a significant impact on stratification. Any additional freshwater influx due to monsoon intensification etc. would tend to increase the stratification even further.

[9] Section 2 presents the Mediterranean model with 2-layer hydraulic control through a realistic triangular shaped Gibraltar strait and sill, whose section changes as the sea level rises. This is the first time this continuously changing strait geometry has been studied. The entire 18,000 years from the Last Glacial Maximum to the present day are modeled. Unlike previous paleo/hydraulic studies [e.g., *Rohling*, 1994; *Rohling and Bryden*, 1994; *Rohling*, 1999], this model is (1) time-dependent (the strait and the salinity of the Atlantic inflow vary), (2) there is explicit feedback between the basin and straits through water formation, and (3) the model represents submaximal exchanges as well as maximal. Section 3 describes three model runs following the time-dependent stratification of the Mediterranean throughout the deglaciation. Section 4 discusses the implications for the stratification and the formation of Sapropel S1, and section 5 concludes.

2. Model Description

[10] The Mediterranean is modeled with three boxes, see Figure 3, all of which can vary in volume. The main boxes are for upper and lower layer water (U and L) which flow in and out at the Gibraltar Strait. The third box is a water formation box F where surface/boundary processes of freshwater and heat exchange are applied. The volume, heat and salt exchanges between the different boxes represent diffusion and convection processes, and the model will conserve all three quantities.

[11] The hydraulic exchange with the Atlantic depends upon the sea level in the Atlantic, the sea level in the basin, the interface depth between upper and lower layers inside the basin, and on the density difference between these layers. The sea surface in the Atlantic, the excess evaporation and heat loss in the basin, and the salinity and temperature of Atlantic waters are specified independently. From these (potentially time varying) input parameters the model calculates the time evolution of the sea level in the Mediterranean, the interface depth in the strait and basin, the salinity difference between the two layers, and the strait flow rates.

2.1. Strait Equations

[12] In the following, the variables are (see also Figures 3 and 4): volume transports q, depths below the reference level h, and section areas across the strait, a. Strait variables are subscripted 1 or 2 according to the layer, whereas basin quantities are subscripted by the Roman letters F, U or L, according to the box. To determine the exchanges at the

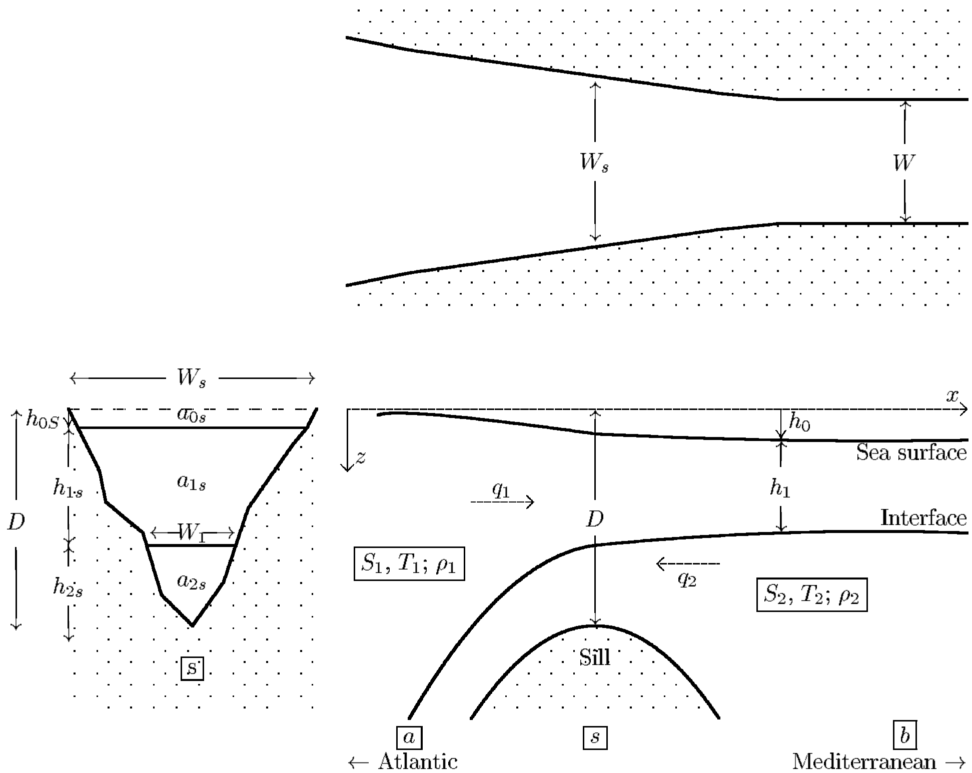

Figure 4. The definition of the variables in the strait in the HYCOBOX model. While the model allows cross sections of arbitrary shape, we use a triangular cross section for this study.

strait at any given time the following set of 6 nondimensional equations is solved iteratively:

$$0 = B + \frac{\rho_1 h_{0a}}{\rho_2 - \rho_1} \tag{1}$$

$$0 = B - \frac{1}{2}\left(\frac{q_1}{a_{1s}}\right)^2 + \frac{\rho_1 h_{0s}}{\rho_2 - \rho_1} \tag{2}$$

$$0 = B - \frac{1}{2}\left(\frac{q_1}{h_{1b}}\right)^2 + \frac{\rho_1 h_{0b}}{\rho_2 - \rho_1} \tag{3}$$

$$0 = \Delta B - \frac{1}{2}\left(\frac{q_1}{h_{1b}}\right)^2 - (h_{1b} + h_{0b}) \tag{4}$$

$$0 = \Delta B - \frac{1}{2}\left[\left(\frac{q_1}{a_{1s}}\right)^2 - \left(\frac{q_2}{(1 - a_{1s} - a_{0s})}\right)^2\right] - (h_{1s} + h_{0s}) \tag{5}$$

$$1 = \left(\frac{q_1^2}{a_{1s}^3} + \frac{q_2^2}{(1 - a_{1s} - a_{0s})^3}\right) W_1. \tag{6}$$

The first three equations represent the Bernoulli function at the surface, in the Atlantic (1, subscript a), at the sill (2, subscript s), and inside the Mediterranean basin (3, subscript b), with B the Bernoulli potential. Equations (4) and (5) represent the Bernoulli potential ΔB at the upper-lower interface inside the basin and above the sill. The last equation (6) is the Froude number condition for a control point at the sill. The implementation allows for both maximal and submaximal exchange in the strait, depending on the interface depth in the basin $h_{U'}$. The maximal regime is reached when the interface in the basin is shallower than a threshold. In this case the strait transport does not depend on the interface depth, and the interface at the strait entrance h_{1b} is set to a constant value (this value for maximal exchange is calculated separately by solving a similar set of equations as (1)–(6) for an extreme in ΔB) as it is decoupled from the basin by a hydraulic jump. In the submaximal regime (deeper interface), the strait transport decreases with deeper interface, and the interface at the strait entrance is determined by the basin value ($h_{1b} = h_{U'}$).

[13] At the sill, the model allows for arbitrary cross sections (see Figure 4) which are implemented as a lookup table relating the layer thicknesses h_{is} with the cross sections a_{is} (i = 1, 2, 3). For this study, a triangular cross section was used that was twice as wide at the sill as at the narrows ($W_s = 2W$).

2.2. Volume Conservation

[14] In the basin submodel the volume budgets for each box are given below, with the total area of the basin A:

$$\frac{dh_F}{dt} = -(E - P) + c_{UF} - c_{FU} - c_{FL} \tag{7}$$

$$\frac{\mathrm{d}h_U}{\mathrm{d}t} = \frac{q_1}{A} - c_{UF} + c_{FU} - c_{UL} + c_{LU} \tag{8}$$

$$\frac{\mathrm{d}h_L}{\mathrm{d}t} = \frac{q_2}{A} + c_{FL} + c_{UL} - c_{LU} \tag{9}$$

$$h_{U'} = h_F + h_U \tag{10}$$

$$h_0 = H - h_F - h_U - h_L. \tag{11}$$

The last two equations make the relationships shown in Figure 3 clear. The constant $H = 2000$ m is the total depth of the basin from the reference level. The parameters c_{XY} are transport rates between the different boxes, and represent different processes in the basin. The transport rates between upper and lower box, c_{UL} and c_{LU}, represent diffusive mixing across the interface (= pycnocline) in the basin:

$$c_{UL} = c_{LU} = \frac{\kappa}{d_{UL}} \qquad \text{mixing,} \tag{12}$$

where $\kappa = 0.5 \times 10^{-5}$ m^{2}s^{-1} is the background diffusion coefficient, and $d_{UL} = 500$ m the effective diffusion length. Although these equal and opposite mixing exchanges do not affect the layer volumes, they do affect the properties, as discussed in the next section.

[15] The transport rate c_{FL} represents the water mass formation and is critical to the results in this study. A physically reasonable parameterization allows the rate to increase with the density difference if the newly formed water is denser than that of the lower layer. It also decreases if the total thickness of the upper layers becomes small.

$$c_{FL} = \begin{cases} \mu \cdot h_{U'} \cdot (\rho_F - \rho_L) & \text{for } \rho_F > \rho_L \\ 0 & \text{for } \rho_F \le \rho_L \end{cases} \tag{13}$$

In the experiments shown here $\mu = 0.4$ m^3 kg^{-1}s^{-1}. No water is brought back directly from the lower layer into the water formation box.

[16] Finally the water formation box F exchanges water with the upper layer box. The exchange c_{UF} contains a mixing term, but also a one way transfer of volume which is parameterized such that the volume of box F is forced back to a specified volume $h_{F\ \mathrm{set}}$ on some relaxation time t_{relax} = 0.1 year. Therefore:

$$c_{UF} = \frac{\kappa}{d_{FU}} + \frac{h_{F\ \mathrm{set}} - h_F}{t_{\mathrm{relax}}} \tag{14}$$

$$c_{FU} = \frac{\kappa}{d_{FU}}. \tag{15}$$

Here, $\kappa = 0.5 \times 10^{-5}$ m^{2}s^{-1} as before, but $d_{FU} = 10$ m, reflecting the smaller vertical length scale near the surface. In most experiments, the specified volume for the water formation box F was assumed to be 30% of the volume of box U, i.e., $h_{F\ \mathrm{set}} = 0.3 \cdot h_U$. This choice is somewhat arbitrary, but the system is not sensitive to the value of $h_{F\ \mathrm{set}}$.

2.3. Salt and Heat Conservation

[17] The transport between the boxes affects the salinity and temperature of each box, as the water is mixed. The water properties of the upper layer box U are also influenced by the inflowing Atlantic water with S_1 and T_1; and the water mass formation box F is subject to boundary fluxes, i.e., excess evaporation $E - P$ and the atmospheric heat loss H_{Atm}. This leads to equations for salt conservation

$$Ah_F \frac{\mathrm{d}S_F}{\mathrm{d}t} = (S_U - S_F) \cdot c_{UF} + S_F \cdot (E - P) \tag{16}$$

$$Ah_U \frac{\mathrm{d}S_U}{\mathrm{d}t} = (S_1 - S_U) \cdot q_1 + (S_L - S_U) \cdot c_{LU} + (S_F - S_U) \cdot c_{FU} \tag{17}$$

$$Ah_L \frac{\mathrm{d}S_L}{\mathrm{d}t} = (S_U - S_L) \cdot c_{UL} + (S_F - S_L) \cdot c_{FL}, \tag{18}$$

and heat conservation

$$Ah_F \frac{\mathrm{d}T_F}{\mathrm{d}t} = (T_U - T_F) \cdot c_{UF} - \frac{H_{\mathrm{Atm}}}{C_{\mathrm{water}}\ \rho_F} \tag{19}$$

$$Ah_U \frac{\mathrm{d}T_U}{\mathrm{d}t} = (T_1 - T_U) \cdot q_1 + (T_L - T_U) \cdot c_{LU} + (T_F - T_U) \cdot c_{FU} \tag{20}$$

$$Ah_L \frac{\mathrm{d}T_L}{\mathrm{d}t} = (T_U - T_L) \cdot c_{UL} + (T_F - T_L) \cdot c_{FL}, \tag{21}$$

where the last terms in (16) and (19) are the changes due to air-sea interaction in the water formation box. C_{water} is the specific heat capacity of seawater. The model exchanges conserve total volume, salt and heat between the layers.

[18] The validity of the model can be checked using conditions similar to the present day. For today's sea level, an evaporation of 75 cm/year, a heat loss of $H_{\mathrm{Atm}} = 7$ W/m^2 over the whole basin, an inflow salinity of $S_1 = 36$ psu and temperature of $T_1 = 16$°C, the model calculates a pycnocline depth of $h_{U'} = 81.8$ m in the steady state, and an outflow of $q_2 = 1.05$ Sv. The model is in the marginally submaximal regime. For comparison, the maximal regime would have a lower layer transport of $q_2 = 1.06$ Sv and a shallower interface depth of $h_{U'} = 74$ m.

[19] Under these conditions the lower layer has a salinity of $S_2 = 37.98$ psu and temperature $T_2 = 12.52$°C thus giving differences with the inflow of $\Delta S = 1.98$ psu and $\Delta T = 3.48$°C. This is in reasonable agreement with observed present-day values. For example, *Bray et al.* [1995] measured a salinity difference between in- and outflow of 2.0 psu and a temperature difference of 3.0°C, averaged over five sections at different times of year. Similar results were obtained by *Baringer and Price* [1997].

Table 1. Sensitivity of the Model to Variations in the Water Formation Parameter μ and the Diffusion Coefficient κ (or Equivalently the Diffusion Length d)[a]

κ	μ	$h_{U'}$, m	q_2, Sv	S_2, psu	T_2,°C
5×10^{-5}	0.4	81.8	1.051	37.98	12.52
4×10^{-5}	0.4	91.7	1.035	38.03	12.44
6×10^{-5}	0.4	71.8	1.057	37.97	12.55
5×10^{-5}	0.32	93.5	1.018	38.05	12.42
5×10^{-5}	0.48	69.3	1.057	37.97	12.54

[a]In the two cases with increased mixing, the system is in the maximal state.

[20] The sensitivity of the basin quantities to variations in the water formation parameter μ and the mixing parameter κ is shown in Table 1, by varying each by ±20%. Both these parameters control mixing between the layers, but with μ operating via the water formation layer where surface fluxes of heat and freshwater also occur. It can be seen that most basin quantities are insensitive to both μ and κ variations, except for the interface depth $h_{U'}$ which shows broadly the same sensitivity to both parameters. This lack of sensitivity is partly a consequence of the fact that the basic state is close to maximal. Increased mixing via either μ or κ drives the basin to maximal exchange (the "overmixed" state) and decreased mixing takes it further from maximal conditions, as shown by changes in $h_{U'}$.

3. Deglaciation Experiments

3.1. Forcing Scenarios

[21] The model was used to investigate the effects of sea-level change. Three experiments were performed, with the model being run for 18 kyr each time. For all runs, the sea-level curve from Figure 2 defines the Atlantic sea level h_{0a}. The Atlantic inflow salinity was assumed to vary with the total global ice volume and was therefore modeled as

$$S_1 = S_{1\text{ present}} \frac{H_{\text{Global ocean}}}{H_{\text{Global ocean}} - h_{0a}}, \tag{22}$$

where the present-day salinity $S_{1\text{ present}}$ = 36 psu and the average depth of the global ocean $H_{\text{Global ocean}}$ = 3800 m. For the inflow temperature, a constant value of T_1 = 16°C was used, and variations during the Holocene were neglected as they are difficult to quantify and have only a minor effect on the results of this paper.

[22] The three experiments have the same surface heat flux of h_{Atm} = 7 W/m^2 but differ in the imposed freshwater flux (see Figure 5 and Table 2). In the first experiment (CONST), the freshwater flux remains constant throughout with a value for the net evaporation chosen to be 75 cm/year, so that stratification changes are purely due to sea-level changes. This $E - P$ should be regarded as a baseline against which to measure the impact of strait and Black Sea inputs. It is at the upper end of present-day estimates but the results would not change drastically if a slightly lower baseline $E - P$ were chosen [see *Matthiesen*, 2001] for a full discussion of the sensitivity of the model to $E - P$).

[23] The second experiment (GRAD) tries to model the gradual opening of the Black Sea, based on the scenario in the work of *Lane-Serff et al.* [1997]: Between 8500 year BP and 5000 year BP, there is an additional influx of fresh water as the fresh Black Lake water is replaced. The influx increases linearly from 8500 year BP, reaching a peak of 15000 m^3/s (equivalent to 20 cm/year reduction in net evaporation over the whole basin) at 6750 year BP, and then decreases until 5000 year BP. The third experiment (CATA) models the catastrophic scenario: Before the opening of the Black Sea, the missing freshwater influx through the Bosphorus means that the excess evaporation would have been 20% larger than today. With the catastrophic event at 7150 year BP, the Black Sea outflow was established. Initially, the outflow would have exceeded present-day values as the fresh Black Lake water was lost, similar to the gradual scenario. In this experiment, the excess evaporation drops from 90 cm/year to 55 cm/year at the opening and then returns linearly to its present-day value of 75 cm/year over the following 1750 years.

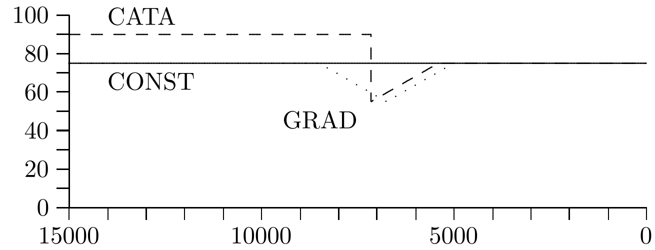

Figure 5. Freshwater budget for the Holocene experiments. Experiment CONST (solid line) uses a constant freshwater budget of $E - P$ = 75 cm/yr, experiment GRAD (dotted line) models the gradual opening of the Black Sea with increases freshwater inflow during the transition, and CATA (dashed line) models the catastrophic opening of the Black Sea after *Ryan et al.* [1997].

3.2. Results

[24] Figure 6 shows the interface depth and the salinity for experiment CONST. This experiment shows the varying basin situation during the deglaciation in the absence of any additional Black Sea freshwater flux. At the beginning of the run (equivalent to the LGM), the Strait is shallower and narrower than today, due to the lower sea level, and the reduced exchange with the Atlantic means that the salinity contrast between the Atlantic and the Mediterranean Sea was higher than today. Also, the Atlantic salinity was higher, so that the lower layer salinity in the basin is 41.4 psu, considerably higher than observed today.

[25] The strait flow at this time was well into the maximal exchange regime. This replicates the results obtained by *Rohling and Bryden* [1994] with a steady state model with a

Table 2. Parameters for the Holocene Experiments

Experiment	Variation in Net Evaporation
CONST	constant
GRAD	8500–6750 BP: linear decrease by 20 cm/year
	6750–5000 BP: linear increase by 20 cm/year
	before 7150 BP: 20% (15 cm/year) higher
	7150 BP: sudden decrease by 35 cm/year
CATA	7150–5400 BP: linear increase by 20 cm/year

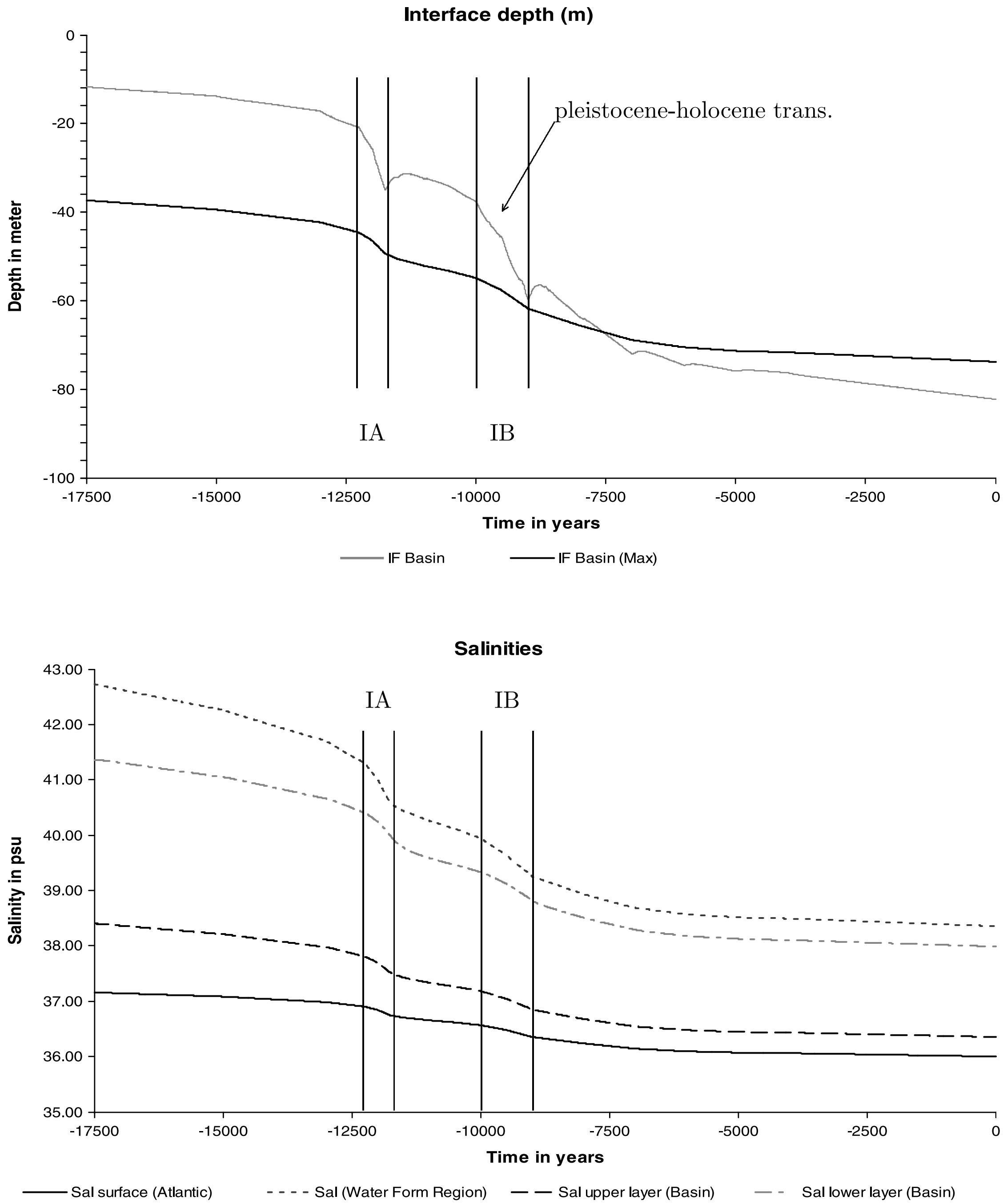

Figure 6. The experiment with constant freshwater budget (CONST). (top) The interface depth in the basin (thin line, "IF Basin") and the switchover depth above which the system is in the maximal regime (thick line, "IF Basin (Max)"). The interface deepens considerably around the pleistocene-holocene transition, indicating reduced deep/intermediate water formation. (bottom) The salinities for the different boxes. The salinity of the lower layer (dash-dotted) can be regarded as the basin average. See color version of this figure in the HTML.

triangular strait, although they used a somewhat smaller value for the net evaporation (56 cm/year).

[26] The residence time is determined by the strait transport and lower layer volume and is of the order of centuries or millennia, so that the conditions in the basin (in particular the lower layer salinity) can lag behind the steady state situation especially during the Meltwater Peaks. As the sea level rises and the Atlantic exchange increases, lowering the surface salinity, the intermediate and deep waters still retain a comparatively high density, leading to stronger stratification in the basin and reduced water formation (Figure 6).

[27] In addition the interface depth deepens suddenly during these events. Of particular interest is the Pleistocene-Holocene transition at 9600 year BP, at which the

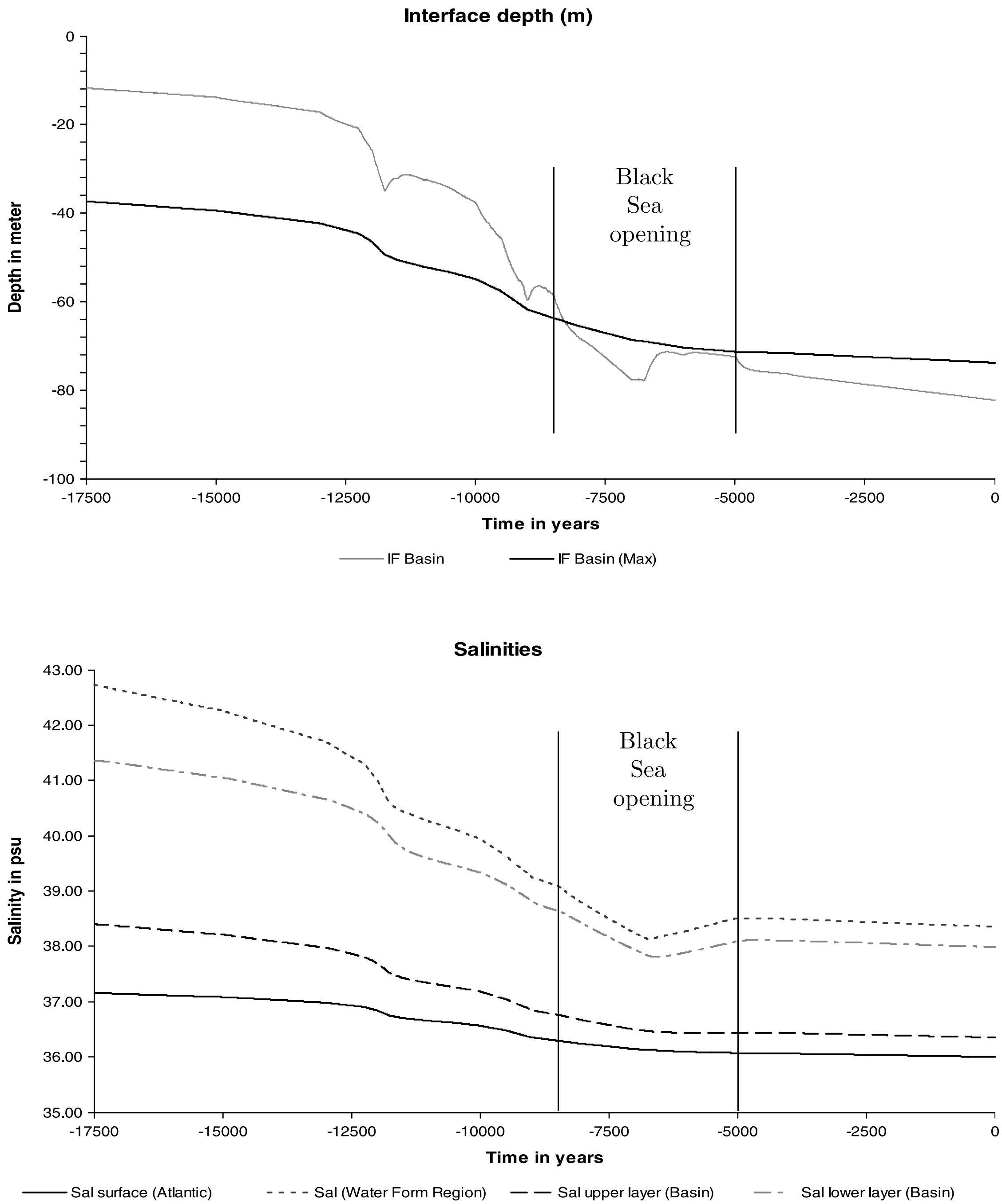

Figure 7. (top) The interface depth and the salinity for the experiment with the gradual opening of the Black Sea (GRAD). (bottom) The additional freshwater influx leads to freshening of the Mediterranean. See color version of this figure in the HTML.

disappearance of planktonic foraminifera of the genus *Neogloboquadrina* is often interpreted as an indication that the pycnocline dropped below the photic zone [*Rohling and Gieskes*, 1989]. In the model the interface depth deepens rapidly, in response to Meltwater Peak 1B, from 38 m at 10000 years BP to 60 m at 9000 year BP. Soon after this second Meltwater Peak, at 7500 year BP, the basin becomes marginally submaximal and remains so to the present day. Notice that at the end of the run the parameters are fairly consistent with present-day Mediterranean measurements. The salinity difference between inflowing Atlantic water and Mediterranean outflow is about 2.0 psu with the salinity in the water formation region being about 0.4 psu higher than that of the deep water. Although the flow is submaximal at the end of the experiment it is only marginally so. We now compare these effects with the impact of additional freshwater from the two scenarios for the opening of the Black Sea.

[28] The experiment for the gradual opening scenario (GRAD), Figure 7, is identical to CONST before the onset of the freshwater influx from the Black Sea. The interface

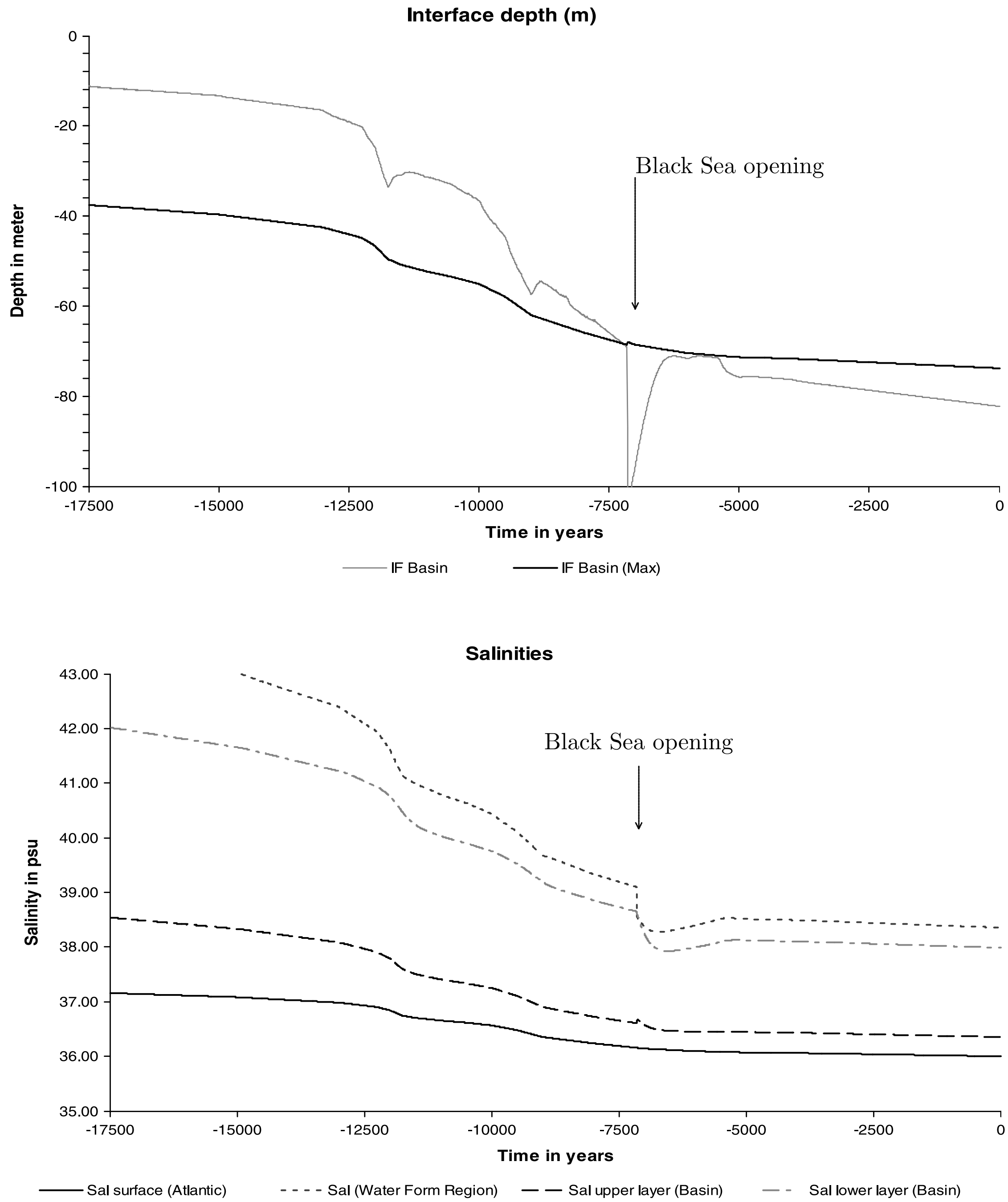

Figure 8. The interface depth and the salinity for the experiment with the catastrophic opening of the Black Sea (CATA). At the opening, the interface drops rapidly, making the strait exchange submaximal (top). See color version of this figure in the HTML.

depth drops more rapidly than in CONST during the first half of the freshwater event, i.e., during the time when the net evaporation decreases (8500–6750 year BP). The strait flow becomes considerably submaximal at this time, but the interface starts to rise again in the second half of the event (6750–5000 year BP), leading to a slightly shallower interface than at this time in CONST. The initial large drop in interface indicates a reduced water formation, and as *Lane-Serff et al.* [1997] noted, this may have been a factor in the formation of sapropels.

[29] The additional freshwater influx from the Black Sea leads to a considerable freshening of the Mediterranean in the 7th millennium BP, evident from the minimum of the salinity in the water formation layer and the lower layer (Figure 7 bottom) centered around 6500 years BP, i.e., 200–300 years after the largest freshwater input. The timing of the event used here is rather late for the sapropel formation (9600–6750 year BP), but there are in any case considerable uncertainties in the timing and scale of the additional freshwater influx.

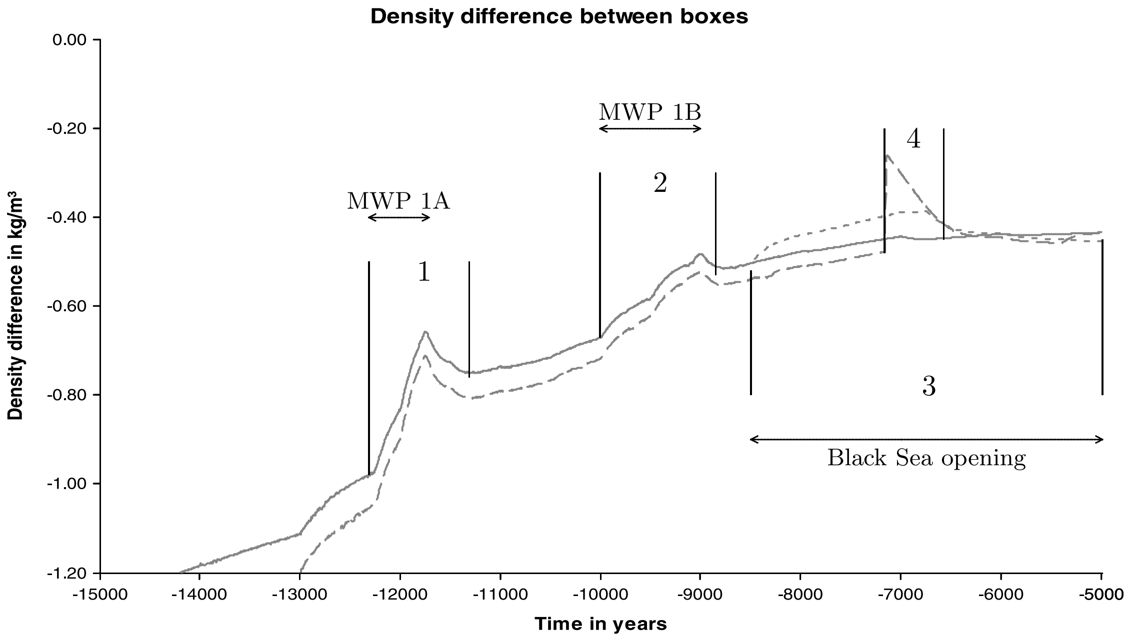

Figure 9. The density difference between boxes F and L for the three experiments. CONST, solid line; GRAD, dotted line; CATA, dashed line. The experiment for the catastrophic opening of the Black Sea shows a stronger stratification before the event as the excess evaporation is higher. Vertical lines mark the beginning and end of the different peaks in stratification, see also Table 3, and the arrows mark the duration of the Meltwater Peaks and the gradual Black Sea opening. The stratification shows local extrema 300–500 years after the freshwater events IA and IB. See color version of this figure in the HTML.

[30] In the experiment CATA (Figure 8), the initial salinity in the basin at 17500 ka BP is approximately 0.6 psu higher than in the other two experiments; and 0.4 psu higher prior to the opening at 7150 years BP. This timing is chosen to coincide with the sediment unconformity noted by *Ryan et al.* [1997]. At the time of the opening, the upper layer and water formation layer salinities show a sharp drop, and the interface depth increases suddenly, well into the submaximal regime for the straits. The drop in interface depth lasts for approximately 500 years with an e-folding time of 200 years. The sudden freshening of the surface layer leads to more stable stratification, and the water formation only recovers as the density of the bottom layer adjusts on a timescale of 200 years. The timescale is therefore mainly determined by the slow changes in salinity in the basin.

[31] While the change in interface depth indicates a reduced circulation for half a millennium after the catastrophic opening of the Black Sea, the timing makes it unlikely that the flooding of the Black Sea and subsequent increase in freshwater influx into the Mediterranean are a causal factor for sapropel formation, which began approximately 2500 years before the event. However, the additional freshwater may have been a factor in maintaining a low circulation toward the end of the deposition of sapropel S1, i.e., from 7150 years BP to 6400 years BP.

[32] Comparing the two scenarios, the catastrophic scenario could lead to a shorter, but more complete collapse of the circulation than the gradual scenario. Timing of the sapropel layer and its duration could be consistent with the gradual scenario, but makes the catastrophic opening an unlikely primary cause for S1.

4. Discussion of Stratification Changes

[33] The longest timescale of adjustment for the Mediterranean basin is that of the salinity of the lower layer water. The large volume of water compared with the small amount of formation each year guarantees that this will lag behind any trends imposed from external conditions such as sea-level rise or changing net surface fluxes. This conclusion can be drawn independently of how water formation is represented in the basin. This lag in the deep water properties (the reservoir effect) has implications for the stratification and hence ventilation of the basin during times of rapid trends in boundary conditions [*Rohling*, 1994].

[34] As the sea level rises, the exchange of water through the Strait of Gibraltar increases. If the surface fluxes do not change this will mean that the surface water properties must become less dense. However, the salinity of the deep and intermediate waters can change only slowly, leading to a transient increase in stratification which is particularly pronounced during the Meltwater Peaks. This reservoir effect is already present in the CONST experiment, and the three model experiments can be used to compare this reservoir effect with the effect of the changing water budgets in the two experiments GRAD and CATA.

[35] The most direct indicator of the circulation regime would be the water formation rate c_{FL}. However, on timescales of centuries or millennia, c_{FL} is not a sensitive

Table 3. Comparison of the Peaks in Stratification in Figure 9

Peak	Age, kyears BP	Peak Width, kyears	$\Delta\rho$, kg/m^3	Baseline, kg/m^3	Peak Height, kg/m^3	Change, %
Meltwater IA	11750	1000	−0.65	−0.83	0.18	21
Meltwater IB	9000	1250	−0.48	−0.56	0.08	14
Gradual	6750	3500	−0.39	−0.45	0.06	13
Catastrophic	7150	600	−0.26	−0.46	0.20	43

parameter in a two-layer basin as it has to balance the strait transport on comparatively short timescales of a few years. The results of these brief changes in water formation are indirectly visible in the movement of the interface depths in Figures 6, 7 and 8.

[36] A better indicator is the density difference $\Delta\rho = \rho_L - \rho_F$ between the lower layer L and the water formation box F. Figure 9 shows this density difference for the three experiments. If this density difference is small (i.e., the graph shows a peak in the positive (upward) direction), the newly formed water is not very dense and therefore less likely to sink, so that the water formation is reduced.

[37] For both scenarios, the curves show peaks at the respective meltwater events, confirming that this variable is a sensitive indicator of the circulation regime. In view of the reservoir mechanism, the curves show clear peaks at the two main meltwater events at 11750 years BP and 9000 years BP, three to five centuries after the strongest meltwater influx (12000 years BP and 9500 years BP). Table 3 shows the different peaks in comparison. In each case, the peak height is calculated from the value compared to a baseline. For the two freshwater events, the baseline is taken from the experiment with constant freshwater budget, which serves as a control run. For the two Meltwater Peaks, no control run is available, and the height of the peak is estimated as the height above a hypothetical straight line connecting across the base of the peaks.

[38] Comparing the two freshwater events, the catastrophic scenario leads to a considerably higher peak of 0.20 kg/m^3 than the gradual scenario (0.06 kg/m^3), i.e., the density difference between the newly formed deep water F and the lower layer is reduced by 43% for the catastrophic opening, and by 13% for the gradual scenario, but the catastrophic scenario lasts only 600 years while the gradual scenario lasts 3500 years. However, Meltwater Peak 1A leads to a stratification peak height of 0.18 kg/m^3, i.e., a reduction of 22%, considerably higher than the gradual scenario. This would indicate that this Meltwater Peak could have a similar or even stronger impact on the circulation as the Black Sea events. Also the later Meltwater Peak 1B leads to a stratification peak that is 0.08 kg/m^3 (14%), more than for the gradual opening of the Black Sea (0.06 kg/m^3). The peaks in stratification due to the meltwater events are sufficiently high to suggest that the reservoir effect would have had a considerable effect on the circulation of the Mediterranean, independent of other freshwater sources.

[39] This result is not sensitive to the choice of the model parameters μ, κ and d_{UL}. Although μ influences the interface depth (see Table 1), it does not affect the densities of the boxes unless it is so small that it leads to a situation well inside the submaximal regime, which is inconsistent with the present situation. The diffusion parameters κ and d_{UL} do influence the densities in Figure 9 (larger mixing reduces the density differences between the boxes), but the relative size of the peaks in Figure 9 and Table 3 remain the same. Therefore we believe that the HYCOBOX result based on the comparison of the scenarios is robust.

[40] The timing of the different events (see Figure 1) suggests that the sapropel layers are deposited following the Meltwater Peak 1B, or during the gradual opening of the Black Sea. The result of these experiments show that this Meltwater Peak had a slightly smaller effect than the gradual opening of the Black Sea. However, the timing of the Black Sea opening is rather late compared to the sapropel deposition, and as the salination of the Black Sea is accurately and independently dated to 7150 years BP, there is only limited scope for an earlier freshwater influx from the Black Sea. Therefore it is fair to speculate that the Meltwater Peak 1B may have influenced the onset of the sapropel formation. Furthermore, if the catastrophic scenario is accepted over the gradual scenario, the Meltwater Peak 1B still remains as a likely cause for reduced circulation during this period.

[41] It is unclear, however, why the much stronger Meltwater Peak 1A did not lead to sapropel formation. Possible explanations include additional changes in the freshwater budget that were not considered here, or stronger cooling in the Younger Dryas, intensifying the deep water formation. It also seems likely that the Mediterranean circulation in general was in a less stable state at the Pleistocene-Holocene transition than during the Late Pleistocene; a question, however, that is beyond the scope of this paper.

5. Summary and Conclusions

[42] This paper quantifies the effects of the changing sea level during the Holocene on the Mediterranean Sea, in particular the reservoir effect: As the sea level rises, the transport through the Strait of Gibraltar increases, making the surface water in the Mediterranean fresher, while the deep water salinity lags behind due to the long residence time, so that the stratification increases, with implications for the circulation. This effect is expected to be most prominent during times of most rapid sea-level change.

[43] A numerical model of a basin-strait-system is used. The basin part of the model consists of three boxes, representing the deep ocean, the surface layer and a water formation layer (i.e., the water masses directly affected by air-sea-fluxes). It is coupled to a strait model based on hydraulic control theory, which allows for both maximal and submaximal strait exchange and has a triangular cross section similar to the cross section at the Camarinal Sill.

[44] The model is run for 18000 years, i.e., from the Last Glacial Maximum (LGM), imposing a rising sea level in the Atlantic based on the sea-level curve by *Fairbanks* [1989].

The sea level rises continuously from 120 m below present-day level, with two periods of faster sea-level rise ("Meltwater Peak 1A" and "1B"). The release of fresh water from the Cryosphere also leads to changes in the salinity of the Atlantic inflow. The effects on the stratification in the Mediterranean during the meltwater events are quantified in comparison with the effects of two opening scenarios of the Bosphorus, which also lead to freshwater influx from the Black Sea.

[45] The results show a decreasing salinity in the basin for all model runs, with 41.4 psu mean salinity at the beginning during the LGM and 38.0 psu at the end in the present day. We find noticeable peaks in the stratification during the periods of rapid sea-level change. The meltwater events lead to a change in stratification of 21% (Meltwater Peak 1A) and 14% (Meltwater Peak 1B). In the model runs with Black Sea opening, there are additional peaks with the gradual opening scenario changing the stratification by 13% and the catastrophic opening by 43%. Therefore we conclude that the rising sea level alone may have had an effect on the stratification of comparable size to the additional freshwater input from the Black Sea. This mechanism should be seriously considered when investigating e.g., sapropel formation. The model can also be used in further studies of closed ocean basin budgets where the exchanges are controlled by two-layer hydraulics.

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